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Climatological Significance of $\delta^{18}O$ in Precipitation and Ice Cores: a Case Study at the Head of the Ürütnqi River, Tien Shan, China

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Climatological significance of δ^{18}O in precipitation and ice cores: a case study at the head of the Ürümqi river, Tien Shan, China

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ABSTRACT. Stable-oxygen-isotope ratios (δ^{18}
collected from the headwaters of the Ürümqi river, Tien Shan, China, were used to test the
relationship between δ^{18}
temporal relationship is found between δ^{18}
monthly averages which remove synoptic-scale influences such as changes in condensation level, condensation temperature and moisture sources (Yao and others, 1996). Linear fits as high as 0.95‰ °C^{-1} for precipitation events and 1.23‰ °C^{-1} for monthly averages are found.
Although the δ^{18}
(~2 km from the precipitation sampling site) decreased dramatically compared to the precipitation samples, the ice-core records of annually averaged δ^{18}
lated with contemporaneous air temperature, especially summer air temperature, at the nearby Daxigou meteorological station. Nevertheless, the relationship between the ice-core δ^{18}O records and contemporaneous air temperature is less significant than that for the precipitation samples due to depositional and post-depositional modification processes, which are highlighted by the successive snow-pit δ^{18}
No. 1. Our results might extend the application of high-altitude and subtropical ice-core
δ^{18} O–T_a

1. INTRODUCTION

The climatological significance of δ^{18}
from the tropics and subtropics has received much less attention than that of δ^{18}O in precipitation from the polar regions. However, pioneering work by Rozanski and others (1992), using the past three decades of δ^{18}
cipitation and surface air temperature available through the International Atomic Energy Agency – World Meteorological Organization global network, indicated that the slope values of δ^{18}O–T_a varied from 0.71 for high-latitude areas to virtually zero in the tropics where a strong relationship between δ^{18}
(Dansgaard, 1964; Grootes and others, 1989). Alternately, a "precipitation-amount effect" for δ^{18}O and δD was also observed both on Himalayan slopes and in the southern/central region of the Qinghai–Tibetan Plateau (Wushiki, 1977a; Wake and Stiévenard, 1995). However, recent work suggests that the δ^{18}
lected on the northern Qinghai–Tibetan Plateau during the period 1991–94 are positively correlated to T_a (Yao and others, 1996), with slopes of linear fit ranging from 0.29‰ to 0.67‰ °C^{-1}.

On mountainous glaciers, there is evidence of seasonal snowmelt resulting directly from the large seasonal temperature range and the intense summer radiation, and a significant part of the annual accumulation of snow is removed by melting and sublimation (e.g. Dunde ice cap in the Qilian Shan, China; Thompson and others, 1988). Thus the seasonal δ^{18}
duced during recrystallization, in the presence of percolating meltwater (Arnason, 1969). Simultaneously, isotopic fractionation occurs, leaving the solid phase enriched relative to the liquid phase (Arnason, 1969; Búason, 1972). Grabczak and others (1983), using the δ^{18}O and δD profiles from two crevasses on temperate glaciers high in the Andes and in the Himalaya, suggested, however, that the short- and long-term isotopic variations can still be observed in the deeper parts of such glaciers. O and precipit

Despite the strong temporal relationship between δ^{18}
precipitation and T_a on the northern Qinghai–Tibetan Plateau, 30 year records of annually averaged δ^{18}
different ice-coring sites in this region are not correlated significantly with the contemporaneous air-temperature records from their corresponding closest meteorological stations (150– O val
200 km away) (Yao and others, 1996). Seasonal distribution of the precipitation, post-depositional modification of the original isotopic signals in precipitation, and elevation differences between the ice-coring sites and their corresponding closest meteorological stations are known to affect the δ^{18}
T_a relationship. Several ice cores recently drilled from the Qinghai–Tibetan Plateau for paleoclimatic reconstruction

(Thompson and others, 1989, 1997) necessitate a better understanding of the $\delta^{18}O$–T_a relationships in precipitation, as well as in glacier ice in this region.

From the beginning of June 1995 to the end of June 1996, precipitation sampling was conducted at the Daxigou meteorological station (DMS; Fig. 1) at the head of the Ürümqi river, Tien Shan, northwest China, at 3545 m a.s.l. From 11 May to 16 June 1996, a series of snow pits (T in Fig. 1) were also sampled at the nearby Ürümqi glacier No. 1 (1.84 km^2 in area) at 4040 m a.s.l. During May 1996, two ice cores 2 m apart were also recovered at the snow-pit sampling site (TS-1 and TS-2 in Fig. 1). These data provide a special opportunity to test the $\delta^{18}O$–T_a relationships in precipitation in the mid-latitude mountainous region, to identify the effect of the post-depositional processes on the ice-core $\delta^{18}O$ records, and to compare the ice-core $\delta^{18}O$ records to the contemporaneous annual surface temperature recorded at the nearby DMS which has been in continuous operation since 1959.

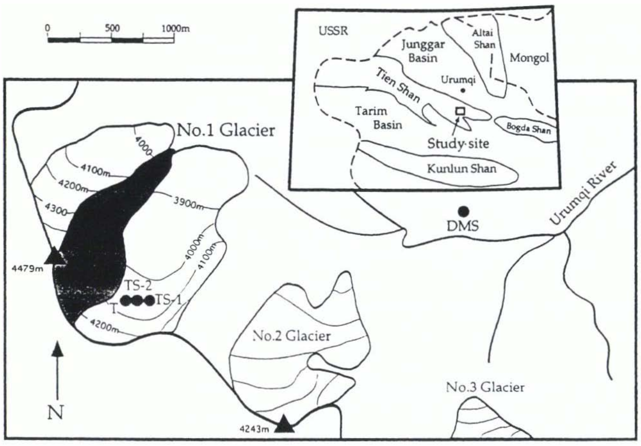

Fig. 1. Sampling sites: DMS, T and TS stand for the Daxigou meteorological station (also the precipitation sampling site), the snow-pit sampling site and the ice-core drilling site, respectively. Inset map shows the location of the Tien Shan in relation to geographic and political features of northwest China.

2. METHODOLOGY

The study site (43.05° N, 86.49° E) is surrounded by large deserts on the north, south and east sides. The nearest sea is located > 3000 km away, so the region is dominated by classic inland climate conditions. The weather conditions in the study area are influenced by the obstacle effect of the Qinghai–Tibetan Plateau. Since the jet stream maintains its west–east orientation along the Tien Shan around the northern part of the Qinghai–Tibetan Plateau (Reiter, 1981), it carries the moisture originating in the Atlantic Ocean and/or the Mediterranean Sea to the study site (Li and Xu, 1984). The annual average surface air temperature and precipitation at the DMS are –5.3°C and 440.6 mm w.e., respectively, for the period 1959–96 (unpublished data).

Precipitation (snowfall and rainfall) samples were collected for each precipitation event during the observation period. To eliminate contamination and sample carryover, identical sampling procedures were used throughout the study (Yao and others, 1996). Plastic containers used for collection were dried (if necessary) and cleaned with a brush between samples. After collection, each sample was placed in a clean plastic bag, melted at about 20°C and poured into pre-cleaned high-density polyethylene bottles, and the tops were sealed in wax to avoid evaporation or diffusion. Bottled samples were transported to the Laboratory of Ice Core and Cold Regions Environment, Chinese Academy of Sciences, and kept in a cold room at –20°C until $\delta^{18}O$ analysis was performed using a Finnigan MAT-252 Spectrometer (precision 0.05‰). Relevant meteorological conditions such as air temperature at the beginning and end of each precipitation event, cloud type, wind speed, wind direction and humidity were recorded for each precipitation sample.

The first snow pit was dug on 11 May 1996, when little ablation of the winter snow had occurred. It was partially refilled after sampling, and sampling on successive days involved digging it out and refacing the sampling surface by at least 30 cm. The same strata were resampled on each occasion after allowing for accumulation or ablation. Snow samples were transferred into pre-cleaned polypropylene bags with plastic scoops for further processing as discussed above.

The ice-core samples were processed in the field by scraping with a clean stainless knife to obtain a contamination-free center sample. Table tops and tools were modified or covered with plastics. The TS-1 and TS-2 ice cores were cut into disks at 3 and 5 cm intervals, respectively. Samples were then transferred into polyethylene bags for further processing as discussed above.

3. RESULTS AND DISCUSSION

3.1. The positive $\delta^{18}O$–T_a relationship in precipitation

The $\delta^{18}O$ values of the precipitation samples show a broad variation; the minimum value of –38.24‰ was obtained on 24 January 1996, and the maximum value of 0.97‰ on 3 August 1995. The arithmetic average value of $\delta^{18}O$ for the samples collected during July and August 1995 is –7.13‰. From 9 July to 17 August 1981, precipitation samples collected at the same sampling site had $\delta^{18}O$ values ranging from –16.0‰ to –1.0‰, with an arithmetic mean average of –7.15‰ (Watanabe and others, 1983). No significant difference is found between these arithmetic averages.

Following Yao and others (1996), we plot the $\delta^{18}O$ and contemporaneous air temperature for individual precipitation events during the period of observation in Figure 2a; the linear relationship between $\delta^{18}O$ and T_a for all individual precipitation events in Figure 2b, and the linear relationship between $\delta^{18}O$ and T_a for individual summer precipitation events (May–October) in Figure 2c. These plots validate the normal assumption that $\delta^{18}O$ is more negative when T_a is cooler and less negative (even positive) when T_a is warmer. The slope of the $\delta^{18}O$–T_a relationship for all precipitation events is 0.95‰ $°C^{-1}$, which is higher than that reported in other mid-latitude to tropical regions (Rozanski and others, 1992; Yao and others, 1996) and in polar regions (Dansgaard and others, 1964; Jouzel and Merlivat, 1984; Mosley-Thompson and others, 1990; Peel, 1992). Since in the Tien Shan, ~90% of the precipitation occurs during summer, we exclude $\delta^{18}O$ values of winter precipitation samples and plot the rest in Figure 2c. The strong relationship between $\delta^{18}O$ and T_a is still apparent, albeit with a lower slope (0.92‰ $°C^{-1}$). These results support the suggestion by Rozanski and others (1992) that the isotopic composition of precipitation seems to be more sensitive to temperature fluctuations in mountainous regions than in low-altitude areas.

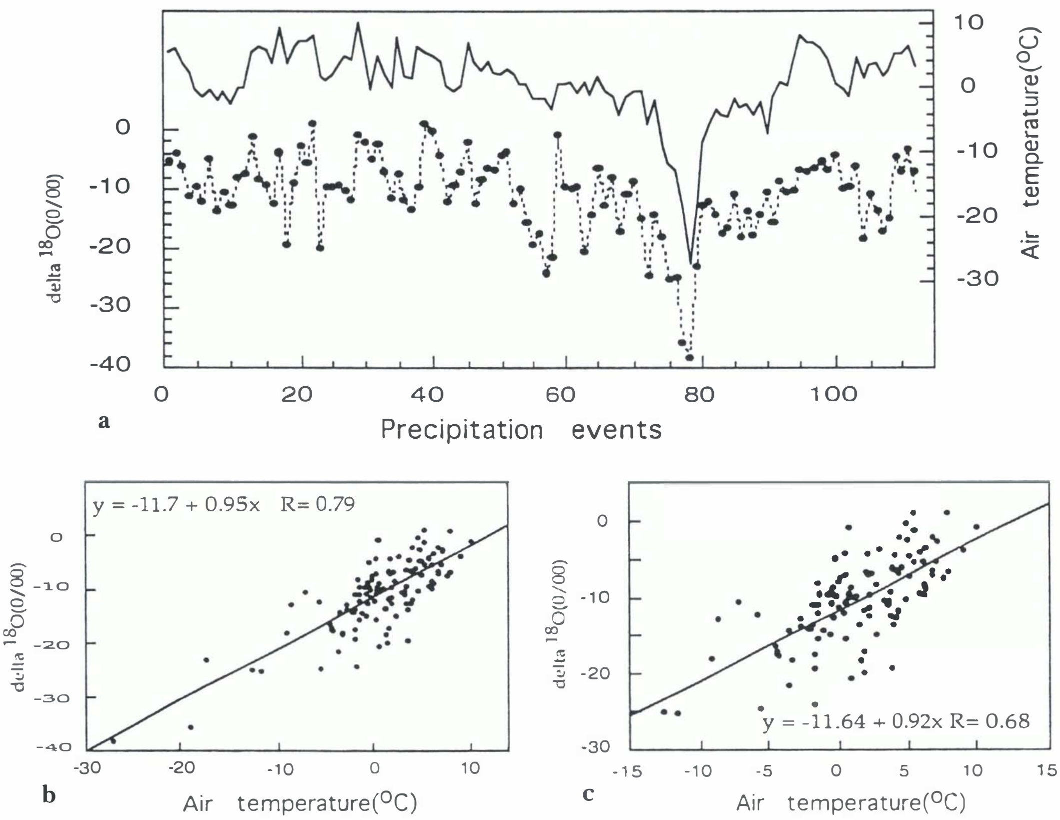

Fig. 2. (a) The relationship between $\delta^{18}O$ and contemporaneous air temperature (T_a) for individual precipitation events. (b) The linear relationship between $\delta^{18}O$ and T_a for all individual precipitation events. The correlation is significant at $p = 0.001$. (c) The linear relationship between $\delta^{18}O$ and T_a for summer individual precipitation events. The correlation is significant at $p = 0.001$.

Although the linear correlation of $\delta^{18}O$–T_a for the individual precipitation events (Fig. 2) is positive, variations in $\delta^{18}O$ do exist that are not attributable to changes in T_a. Other factors, such as the different vapor sources (not significant in our study site), different transport patterns of vapor in the atmosphere, average "rain-out history" of the air mass, and differing temperature structures which control the condensation temperature (Gedzelman and Lawrence, 1982; Covey and Haagenson, 1984), can affect the $\delta^{18}O$–T_a relationships to different degrees (Gedzelman and Lawrence, 1982; Covey and Haagenson, 1984; Fisher and Alt, 1985; Rozanski and others, 1992). Moreover, in the interior of continents, e.g. at our precipitation sampling site, the re-evaporated moisture plays an important role in the atmospheric water balance (Ingraham and Taylor, 1986). Another potentially important influence on the $\delta^{18}O$–T_a relationship is the amount effect (Dansgaard, 1964; Wushiki, 1977a; Wake and Stiévenard, 1995), which is more common in tropical regions and less important in mid-latitudes (Grootes and others, 1989; Rozanski and others, 1992).

As suggested by Jouzel and others (1987), a stronger $\delta^{18}O$–T_a relationship can be expected when individual precipitation events are aggregated into monthly averages, which minimizes the influence of synoptic-scale differences. Figure 3 shows the arithmetic monthly-averaged $\delta^{18}O$ plotted against the monthly-averaged air temperature. The slopes of linear fit are 1.23‰ °C^{-1} for all months (Fig. 3a) and 1.19‰ °C^{-1} for summer months (Fig. 3b). The coefficients (R) for the $\delta^{18}O - T_a$ relationship of monthly averages are much higher than the R values for the individual precipitation events. Undoubtedly, the substantial reduction in the number of data points (degrees of freedom) results in higher errors for the slopes. As suggested by Rozanski and others (1992), the long-term $\delta^{18}O$–T_a relationship for a given

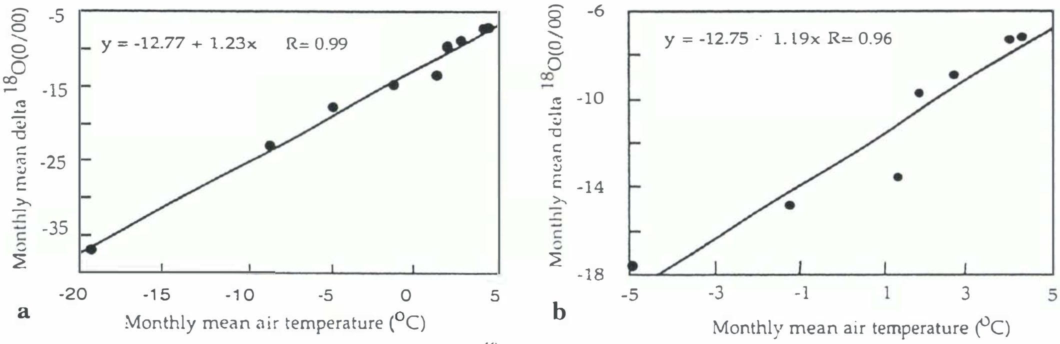

Fig. 3. The linear relationship between monthly-averaged $\delta^{18}O$ and monthly-averaged air temperature T_a: (a) annual; (b) summer. Both are significant at $p = 0.001$. No precipitation samples were collected during November–December 1995 and February–March 1996.

location is believed to be the most appropriate for paleoclimatic reconstruction, because it covers a wide range of different "climate".

3.2. The ice core $\delta^{18}O$–T_a relationship

The ice cores were dated using a variety of different seasonal indicators (e.g. stable isotopes, major anions and cations) from comparison between the TS-1 and TS-2 ice-core records, and comparison with the mass-balance data of the Ürümqi glacier No.1 (Chinese Academy of Sciences, 1982–95). The final dating results are 22 years for the TS-1 ice core and 24 years for the TS-2 ice core (see Fig. 4).

By tritium analysis of the ice samples from a crevasse wall at the head of Khumbu Glacier, Mount Everest, Miller and others (1965) concluded that two strata are deposited in a single accumulation year. One segment represents accumulation during the summer monsoon, and the other accumulation during the months of prevailing winter storms. Wushiki (1977b) also suggested such a conclusion based on the deuterium-content profile of ice cliffs on Kongma Glacier, Khumbu. This stratigraphic characteristic is also apparent in our $\delta^{18}O$ profiles of the TS-1 ice core (Fig. 4), which is supposed to be formed at the beginning and end of each ablation season, when the firnification process is controlled by the thaw–freeze cycles.

Figure 4 shows the $\delta^{18}O$ profiles of the TS-1 and TS-2 ice cores, together with the annual and summer air temperature recorded at the nearby DMS. The $\delta^{18}O$ values of the ice-core samples range from −12.57‰ to −7.74‰, with an arithmetic average of −10.1‰. Compared to the precipitation samples, the $\delta^{18}O$ amplitude in ice cores decreases significantly. However, the distinct annual to seasonal variability of $\delta^{18}O$ values can still be preserved in a fixed annual ice layer. Factors that may contribute to the preservation of the depositional variability are the presence of numerous ice layers in the snowpack, acting as physical obstacles against meltwater percolation, and thick superimposed ice layers that may prevent further smoothing of the $\delta^{18}O$ seasonal signal in underlying ice layers.

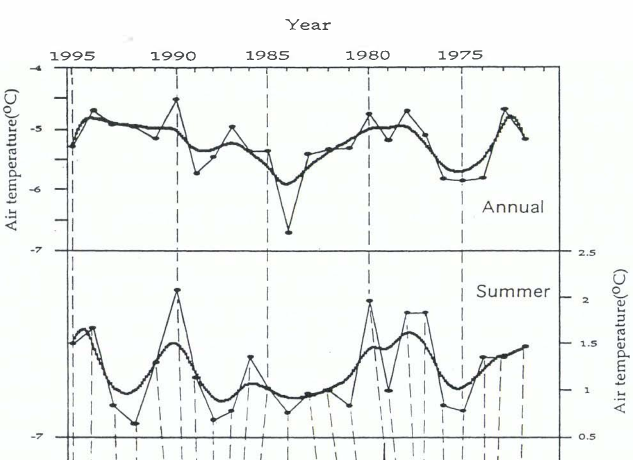

Fig. 4. $\delta^{18}O$ profiles for the TS-1 and TS-2 ice cores, compared with corresponding annual and summer surface air temperature as measured at the DMS. The four coarse solid lines show the smoothing trend using Gaussian weighting coefficients, approximately equal to 5 year moving average, and the gray dashed lines indicate annual ice layers.

Although the TS-1 and TS-2 ice cores were cut at different intervals, similar $\delta^{18}O$ profile features, especially for the smoothing lines, are apparent. The ice-core $\delta^{18}O$ smoothing lines also follow the air-temperature records, particularly for the summer (May–October). As ~90% of the precipitation falls in the summer-half of the year, and $\delta^{18}O$ is only recorded during precipitation, it is reasonable to expect that the $\delta^{18}O$ from an annual ice-core layer records the summer average air temperature of the corresponding year. For further $\delta^{18}O$–T_a comparison, we calculate the mean annual $\delta^{18}O$ value for each year by averaging all sample $\delta^{18}O$ values between two adjacent annual $\delta^{18}O$ peaks, and plot the annually averaged $\delta^{18}O$ against the contemporaneous annual, summer and winter surface temperature (Fig. 5). Apparently, both the TS-1 and TS-2 ice-core $\delta^{18}O$ records might provide a proxy of the summer air temperature, because the linear slopes against the summer temperature are not only higher than that against the annual or winter temperature, but also very similar (0.38 for the TS-1 ice core, and 0.40 for the TS-2 ice core). We notice that the significance of the regression coefficients is far from satisfactory. However, they at least show some progress compared to the results of Yao and others (1996) where the R^2 for annually averaged $\delta^{18}O$ from the ice cores vs surface air temperatures at the closest meteorological station is only in the range 0.000–0.011. Thompson and others (1993) demonstrated that the 5 year running mean of the annual $\delta^{18}O$ averages from the Dunde ice cap (one of Yao and others' three ice cores) was strongly correlated ($R^2 = 0.25$, significance 99.9%) with the 5 year running mean of the Northern Hemisphere annual temperatures from 1895 to 1985. However, such work is not expected here due to the limitation of our short ice-core records. We propose to drill another relatively long ice core from Ürümqi glacier No. 1, and wish to press this question further.

3.3. Depositional and post-depositional modification of ice-core $\delta^{18}O$ records

The relationship between the ice-core $\delta^{18}O$ records and contemporaneous air temperature is less significant than that for the precipitation samples due to depositional and post-depositional processes.

As suggested by Yao and others (1996), the air-temperature data reflect an equal weighting of monthly temperatures, while the ice-core $\delta^{18}O$ record is skewed toward wet-season precipitation. It is unrealistic to expect the $\delta^{18}O$ from an annual layer in ice cores to record the annual average air temperature of the corresponding year. This fact may account for the different linear-fit slopes and correlation significance between contemporaneous annual, summer and winter surface temperature records and the annually averaged ice-core $\delta^{18}O$ records.

The isotopic signal in snowpack may be further modified by post-depositional processes. In polar snow and firn,

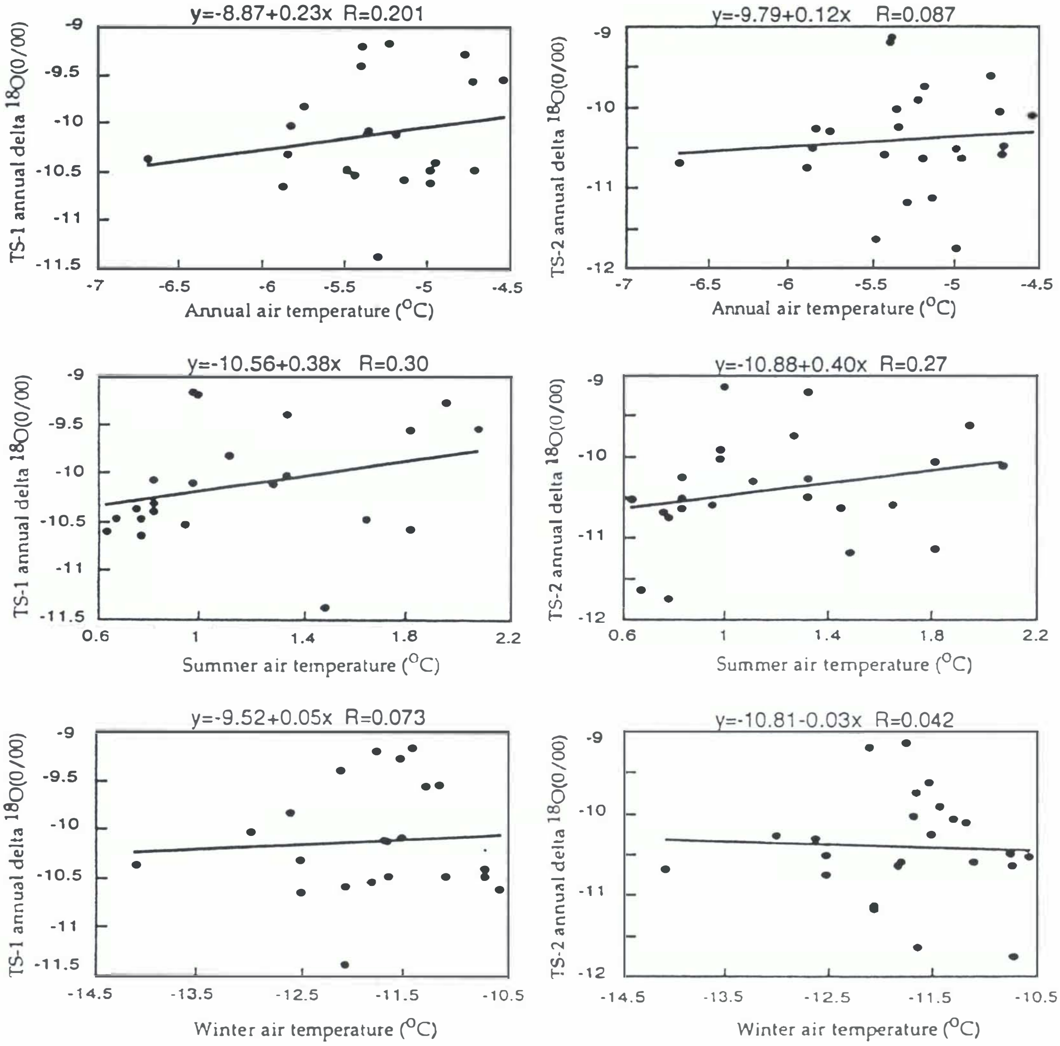

Fig. 5. The linear relationship between annually averaged $\delta^{18}O$ and contemporaneous annual, summer and winter surface temperature.

the isotopic homogenization connects to recrystallization of the grains via the vapor phase, e.g. storms and barometric-pressure changes cause vertical air movements, particularly in the upper firn, where mass exchange between the strata is further accentuated by high temperature gradients (Dansgaard and others, 1973). Such processes can, however, also happen in mountainous glacier snow and firn, especially during the winter-half of the year. Diffusion in the vapor phase also causes considerable interstratificial mass exchange (Dansgaard and others, 1973). In low-accumulation areas, the seasonal $\delta^{18}O$ oscillations are simply missing due to redistribution by snowdrifting or lack of winter (summer) snow, which introduces small-scale and local depositional noise (Fisher and others, 1983). As shown in Figures 6 and 7, the T1 snow pit was sampled at the ice-core drilling site on 11 May 1996 when no snowmelt was observed in the snow-pit stratigraphic profile, and the minimum $\delta^{18}O$ value in the snow pit is −23.29‰, which is much higher than the $\delta^{18}O$ values observed in the winter precipitation samples (e.g. −35.85‰ on 10 January 1996, and −38.24‰ on 24 January 1996). Therefore, the effect of wind-scouring has also been observed on an ice cap, Ellesmere Island, Canada, where the annual mean $\delta^{18}O$ at an ice divide was 2.5‰ less negative than 1.2 km downslope (Fisher and others, 1983).

The effect of snowmelt-percolation processes on the ice-core $\delta^{18}O$ records may be studied using a set of successive snow-pit stratigraphic and $\delta^{18}O$ profiles (Figs 6 and 7). Because the surface of the snow pits changed with time, we adjust the bottoms of all the snow pits to the same 95 cm depth. The T1 snow pit was sampled at intervals of exactly 10 cm. Each $\delta^{18}O$ value presents the average of its corresponding depth range, which covers different stratigraphic snow and firn layers. The minimum $\delta^{18}O$ value of −23.29‰ is present in the middle of the T1 snow pit, which was accumulated during the winter season. The relatively high $\delta^{18}O$ values in the upper and bottom layers correspond to the accumulation deposited during the spring and previous autumn seasons. It is apparent that the $\delta^{18}O$ profile of the T1 snow pit resembles the temperature history during the accumulation period.

The T2 snow pit was dug 3 days later and sampled at its stratigraphic layers. The measured snowpack temperatures were below −2.5°C. Disregarding the upper 15 cm new snow layer, the $\delta^{18}O$ values of the T2 snow pit swing around those of the T1 snow pit. During the period of the T2 and T3 sampling dates, no precipitation event occurred. Though the measured temperatures in the upper layers of the T3 snow pit exceed 0°C at noon, the snowpack below 70 cm remained negative, and several percolation ice layers (or ice lenses) had formed. However, the $\delta^{18}O$ profile of the T3 snow pit preserved the original isotopic characteristics.

Thick superimposed ice layers formed at the bottoms of the T4 and T5 snow pits, and the $\delta^{18}O$ amplitudes decreased dramatically compared with that of the former snow pits,

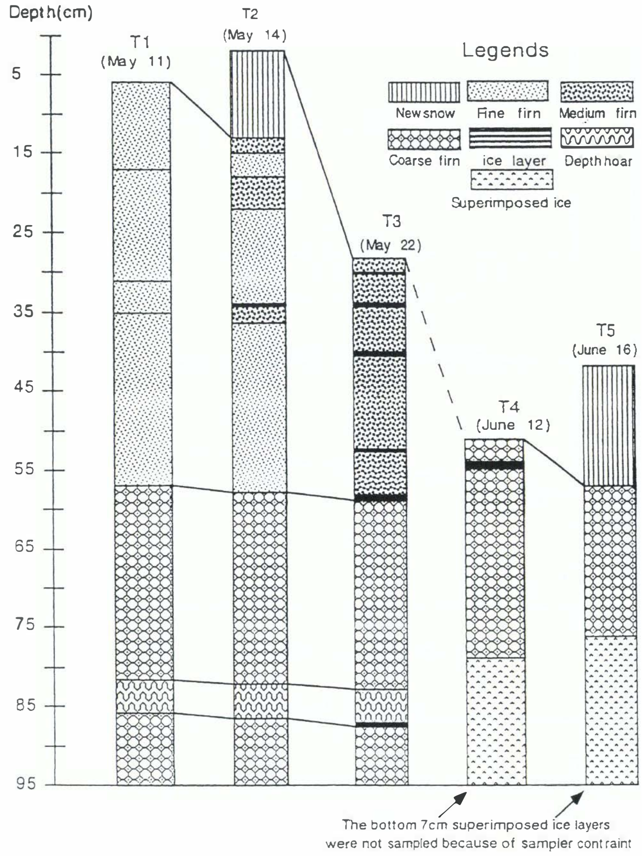

Fig. 6. The successive stratigraphic profiles of snow pits T1 to T5 that were collected during the early-summer melt period.

especially in the superimposed ice layers where the $\delta^{18}O$ values range only from −14.84‰ to −12.57‰. These values are close to the smallest $\delta^{18}O$ (−12.57‰) as observed in the ice cores. Because of the rapid formation of the superimposed ice layer, snowmelt cannot penetrate into the underlying annual ice layers. Thus the modification of the snowpack $\delta^{18}O$ variations occurs only within a certain annual layer.

Koerner (1997) concluded that early summer melt will cause melting of the very negative (cold) $-\delta$ winter/spring snow which lies near the surface. The meltwater then percolates down to refreeze within, or at the base of, the current annual snowpack. Further melt may cause runoff of less negative (warmer) $-\delta$ snow deposited during the early winter/fall period. A very negative (cold) $-\delta$ snowpack remains (Fig. 7). However, the remaining very negative (cold) $-\delta$ snowpack does not necessarily correspond to the winter precipitation with minimum δ values, but rather to a mixture of snow deposited during the winter half of the year. In addition, similar melting and percolation processes can be expected to happen year by year. Isotopic enrichment by partial melting of the snow cover can be roughly estimated by assuming a Rayleigh-type removal of meltwater from the system (Moser and Stichler, 1980). Nevertheless, such isotopic enrichment should influence the ice-core $\delta^{18}O$–T_a records, in such a way that the final isotopic composition of the glacier ice is shifted in the same direction (Grabczak and others, 1983), or even enhance the climatic significance of ice-core $\delta^{18}O$ records, because more enrichment in heavy isotopic content accompanies higher ablation, that is, generally due to higher air temperatures. This guarantees the climatic significance of ice-core $\delta^{18}O$ records from areas of very heavy melt, especially on a long time-scale.

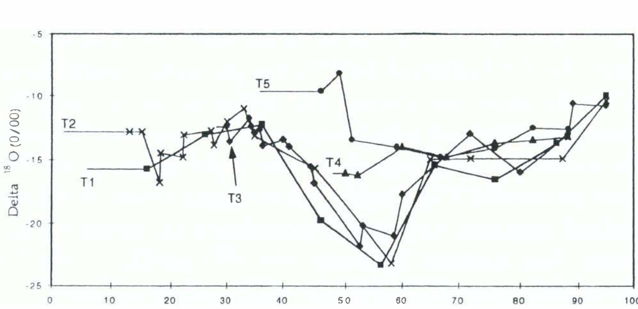

Fig. 7. The successive $\delta^{18}O$ profiles of snow pits T1 to T5 that were collected during the early-summer melt period.

4. CONCLUSIONS

Our data provide the first observations of the relationship between the contemporaneous temperature and $\delta^{18}O$ in precipitation and ice-core samples from the Tien Shan. Although post-depositional processes certainly decrease the significance of the relationship between $\delta^{18}O$ of the precipitation samples and contemporaneous temperature, a positive $\delta^{18}O$–T_a relationship is still observed in ice cores collected from the nearby Ürümqi glacier No. 1. The presence of numerous ice layers in the snowpack and rapid formation of the superimposed ice layer guarantee preservation of the seasonal $\delta^{18}O$ variation in ice cores despite the dramatic decrease in amplitude. Other factors, such as the seasonal distribution of the annual precipitation, isotope diffusion in the snow and firn layers, snowpack redistribution caused by snowdrifting, and the heavy-isotope-content enrichment caused by the partial melting of the snow and firn layers, have a potential but perhaps less important influence on the ice-core $\delta^{18}O$ record. Nevertheless, the climatic significance of mountainous ice-core $\delta^{18}O$ records can be preserved in temperate glacier ice, which might demonstrate that the ice-core climate records can be obtained from less than ideal environments, thus expanding the number of sites in the world where such climatic records could be collected.

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